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Earlier onset of the Indian monsoon in the late twentieth century: The role of anthropogenic aerosols

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[1] The impact of the late twentieth century increase of anthropogenic aerosols on the Indian monsoon onset was investigated with a state-of-the-art climate model with fully interactive aerosols and chemistry. We find that aerosols are likely responsible for the observed earlier onset, resulting in enhanced June precipitation over most of India. This shift is preceded by strong aerosol forcing over the Bay of Bengal and Indochina, mostly attributable to the direct effect, resulting in increased atmospheric stability that inhibits the monsoon migration in May. The adjusted atmospheric circulation leads to thermodynamic changes over the northwestern continental region, including increased surface temperature and near-surface moist static energy, which support a stronger June flow and, facilitated by a relative warming of the Indian Ocean, a vigorous northwestward precipitation shift. These findings underscore the importance of dynamical feedbacks and of regional land-surface processes for the aerosol-monsoon link.
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1. Introduction

[2] Detection and attribution of long-term variations of the South Asian monsoon is of extreme importance. Even small changes in the onset, duration, and spatial distribution of monsoon precipitation (up to 80% of the annual rainfall for most of India) may severely impact highly vulnerable regions (e.g., northwestern India, Pakistan) and strongly affect a substantial fraction of the world's population.

[3] In the past decades, emissions of aerosols over South Asia dramatically increased due to rapid urbanization and population growth. Atmospheric aerosols have the potential to affect precipitation by modulating radiation in the atmosphere through either scattering or absorption (the direct effect) and by altering cloud microphysical processes (the indirect effects). A number of studies recently indicated that increased anthropogenic aerosol over South Asia may have a strong impact on monsoon rainfall [e.g., *Menon et al.*, 2002; *Ramanathan et al.*, 2005; *Lau et al.*, 2006; *Meehl et al.*, 2008; *Wang et al.*, 2009; *Bollasina et al.*, 2011; *Cowan and Cai*, 2011; *Ganguly et al.*, 2012], including the observed drying in the late twentieth century [e.g., *Bollasina et al.*, 2011]. However, the details of the physical pathway for the aerosol-monsoon interaction are still uncertain and debated.

[4] The majority of the studies focused on variations in the summer mean climate only (e.g., the June–September average), while subseasonal changes have received much less attention. Possible changes at the time of the onset could be of equal importance for the agriculture-based monsoon region, as the onset heralds the beginning of the rainy season over the dry Indian subcontinent. *Lau and Kim* [2010] attributed the increase in May–June precipitation over northwestern India to the effect of enhanced upper tropospheric atmospheric heating by absorbing aerosols. *Kajikawa et al.* [2012] argued that enhanced land warming was responsible for the recent earlier shift of the onset over the Bay of Bengal. *Bollasina et al.* [2008] proposed land surface heating by aerosol-induced cloud reduction in May to play a fundamental role in driving the monsoon front inland. *Meehl et al.* [2008] found black carbon aerosols to enhance premonsoon precipitation by increased shortwave absorption and stronger tropospheric meridional temperature gradient. *Wang et al.* [2009] highlighted the role of subcloud moist static energy, initially perturbed by absorbing aerosol heating, in driving a stronger May–June migration of convection over India. On the contrary, *Mahmood and Li* [2013a] found black carbon aerosols to delay the monsoon onset as a result of dynamically induced cooling south of the Tibetan Plateau.

[5] Overall, the nature of changes in the monsoon onset is still an underexplored topic, especially in the context of recent observed long-term monsoon variations and in account of various anthropogenic as well as natural forcings, including a broader perspective on aerosols beyond black carbon alone. Additionally, the physical mechanism behind these changes remains quite elusive and made even murkier by the complex nature of the onset itself. Current coupled climate models, equipped with advanced physics and chemistry, allow for a more mechanistic understanding of the role of aerosols and associated physical mechanisms compared to earlier modeling studies with atmospheric-only models or simplified aerosol physics.

[6] We used a series of historical (1860–2005) experiments with a state-of-the-art climate model, the U.S. National Oceanic and Atmospheric Administration (NOAA) Geophysical Fluid Dynamics Laboratory CM3 model [*Donner et al.*, 2011], to investigate the contribution of

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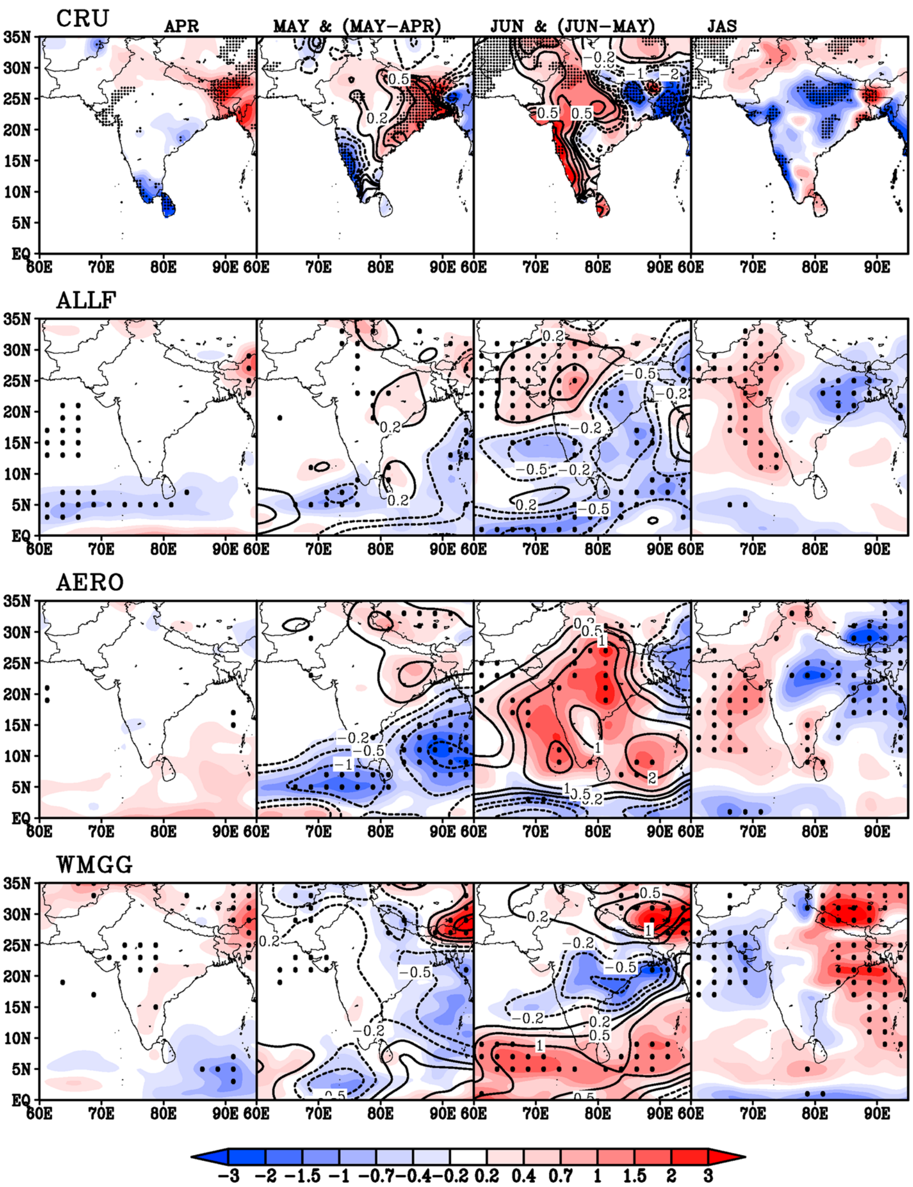

Figure 1. Spatial pattern of the 1950–1999 precipitation trend (mm d^{-1} (50 yr)$^{-1}$, shaded) for (top to bottom) observations (CRU), the all-forcing ensemble (ALLF), the aerosol-only forcing ensemble (AERO), and the greenhouse gases and ozone-only forcing ensemble (WMGG), during (left to right) April, May, June, and the July–September average. The incremental trend (i.e., the difference in the trend with respect to the previous month) is also shown as black contours. The black dots mark the grid points for which the monthly trend exceeds the 95% confidence level according to the two-tailed Student's *t* test.

anthropogenic aerosols to the 1950–1999 observed changes in the Indian monsoon during its development phase. The model has fully interactive aerosols and chemistry and a representation of both direct and indirect aerosol effects. The main simulations analyzed are a five-member ensemble all-forcing experiment (ALLF), with natural and anthropogenic forcings, and 2 three-member ensemble experiments forced only by either anthropogenic aerosols (AERO) or greenhouse gases and ozone (WMGG). Observed precipitation data mainly came from the Climate Research Unit (CRU) TS 3.0 data set [*Mitchell and Jones*, 2005]. The analysis used mainly monthly data; trends were calculated using a linear least squares fit, and their statistical significance was assessed by a two-tailed Student's *t* test accounting for temporal autocorrelation. Incremental trends (i.e., the difference of the trends between two consecutive months) are also displayed to provide a more dynamical portrait of the changes with respect to the underlying unfolding annual cycle.

2. Long-Term Shift in the Monsoon Onset

[7] Observations (CRU in Figure 1) show that from April to June, an area of increased precipitation, initially over northeastern India (where most of the April climatological precipitation occurs), widened and progressively moved westward, replaced by precipitation decrease to the east. In stark contrast, precipitation substantially reduced during the

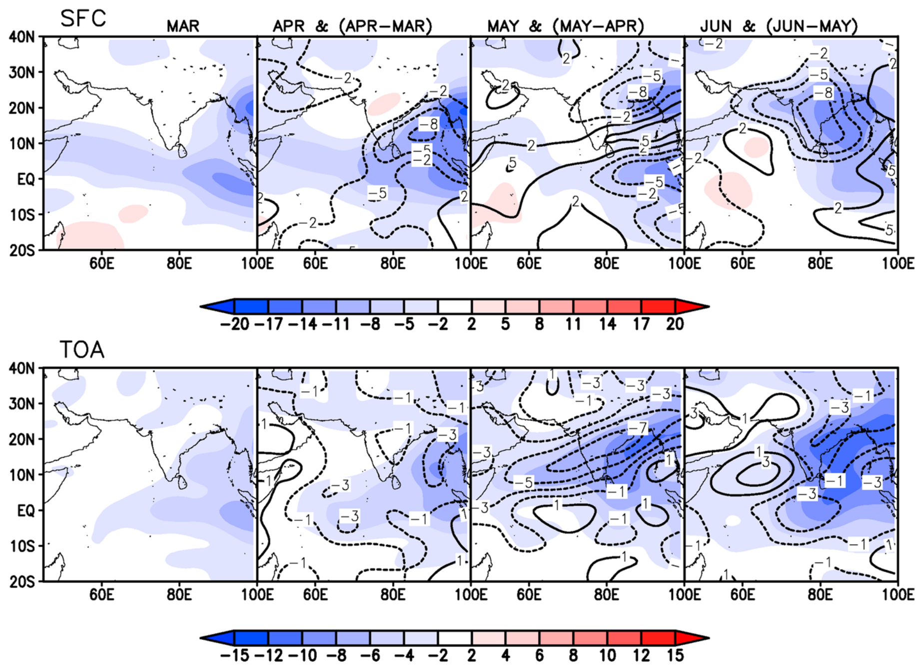

Figure 2. Radiative forcing (W m^{-2}, shaded) of anthropogenic aerosols at the (top) surface and (bottom) top of the atmosphere in the AERO ensemble for March, April, May, and June. The forcing is computed for the period 1940–1999 to account for a possible delay in the climate response in the transient experiments. The incremental trend is also shown as black contours.

rest of summer. These variations are robust among various observational data sets (Figure S1 in the supporting information) and show that the overall summer mean drying is actually the result of two opposing trends between early summer and the rest of the season. The May–June positive trend over central-northern India indicates an earlier monsoon onset; this is further supported by the daily APHRODITE data set which shows the onset date to have shifted at least 2–4 pentads (10–20 days) earlier over the same region (Figure S2).

[8] The all-forcing ensemble (ALLF in Figure 1) reproduces the observed spatial pattern and monthly evolution of the trends reasonably well, especially the gradual northwestward migration of the wetter area and the formation of the summertime east-west dry-wet asymmetric pattern [*Bollasina et al.*, 2011; *Konwar et al.*, 2012] with increased precipitation over Pakistan. Because precipitation is not yet organized and widespread, the relatively low model resolution (~200 km) is likely to play an important role in the underestimation of the observed May precipitation increase over India.

[9] Aerosols are found to be mostly responsible for the earlier monsoon onset (AERO in Figure 1). Two major anomalies are particularly noticeable: a pronounced drying over the north-equatorial Indian Ocean, Bay of Bengal, and Burma in May, and the widespread substantial wettening in June. These two prominent features are clearly dominating the total anthropogenic footprint (ANTHRO in Figure S3) and also emerge in the ALLF changes. Note that the magnitudes of the AERO trends are comparable to the CRU ones, indicating the prominence of the aerosol forcing in driving the overall changes. Forced by greenhouse gases and ozone (WMGG), the model reduces precipitation over central and northern India in May and June. The contribution of natural variability is not completely ruled out, though natural forcings, at least in this model, produce much smaller, if not insignificant, trends (NAT in Figure S3). Interestingly, the aerosol direct effect appears to play the major role in May and June, as precipitation changes associated with the indirect effect only (AEROI in Figure S3) are inconsistent (May) or weak (June).

3. The Mechanism for the Aerosol Impact

[10] The springtime radiative forcing due to increased anthropogenic aerosols is represented in Figure 2. The forcing is computed as the difference in the net radiation trends between two experiments with the atmospheric component of CM3 (i.e., AM3) forced with the same time-evolving historical sea surface temperatures (SSTs) and either (a) historical or (b) preindustrial aerosol emissions. The difference of (a) and (b) provides an estimate of the "adjusted" aerosol forcing, accounting for the fast adjustment of the land-atmosphere system [e.g., *Ming and Ramaswamy*, 2012]. Although the impact of the forcing defined as above could spread outside its boundary through circulation changes, Figure 3 shows that in early spring the largest (negative) forcing both at the surface and at the top of the atmosphere is located over Indochina and the equatorial Indian Ocean, where the largest negative precipitation change occurs (Figure 1). In May, the forcing extends westward across India and increases, with maximum surface values exceeding -16 W m^{-2} over the Bay of Bengal. The forcing further increases in June over India.

[11] The gross features of the patterns displayed in Figure 2 are well correlated with the simulated aerosol optical depth (AOD, Figure S4). The trend in both scattering and absorbing

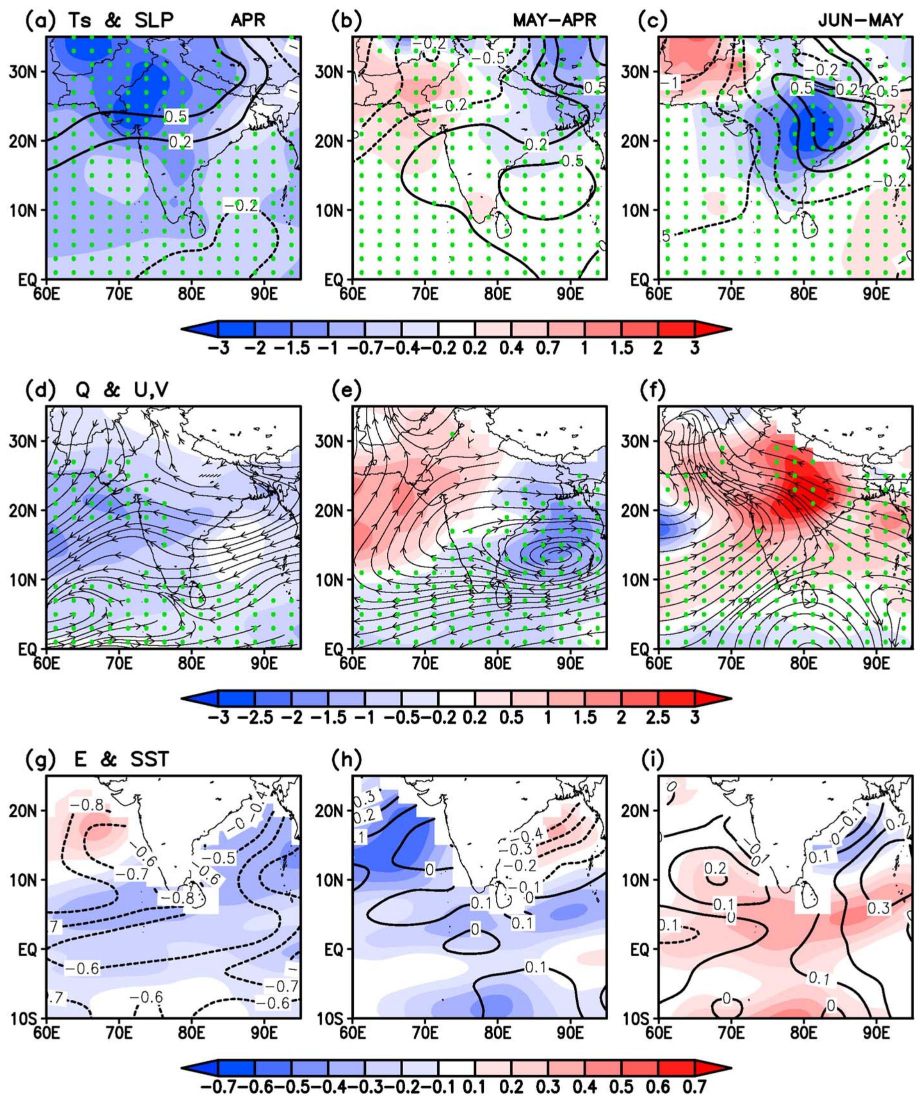

Figure 3. Spatial pattern of the 1950–1999 trends in the AERO ensemble for (left column) April, (middle column) the difference of May and April, and (right column) the difference of June and May for (a–c) surface temperature (K $(50\,yr)^{-1}$, shaded) and sea level pressure (hPa $(50\,yr)^{-1}$, contours), (d–f) 1000–775 hPa integrated specific humidity (kg m^{-2} $(50\,yr)^{-1}$, shaded) and 850 hPa wind (streamlines), and (g–i) evaporation (mm d^{-1} $(50\,yr)^{-1}$, shaded) and sea surface temperature (K $(50\,yr)^{-1}$, contours). The green dots mark the grid points for which the monthly trend exceeds the 95% confidence level according to two-tailed Student's *t* test for surface temperature in Figures 3a–3c and specific humidity in Figures 3d–3f.

AOD is positive almost everywhere and for every month; largest changes appear over Indochina and Burma in spring and over India in June. Notable is the almost incessant early-spring AOD increase over the Bay of Bengal since the late 1940s, mainly associated with increased emissions from premonsoon intense biomass burning [e.g., *Zhao et al*., 2008; *Gautam et al*., 2013] and westward transport by the climatological easterlies. By directly scattering and absorbing incoming shortwave radiation, aerosols cause an enhanced cooling of the surface compared to the layers above and lead to increased atmospheric stability (Figure S5), resulting in reduced precipitation in May. Their indirect effects, as mentioned above, do not appear to be influential in spring. Both scattering and absorption are effective in reducing the net clear-sky surface shortwave radiation (−3.9 and −4.4 W m^{-2} $(50\,yr)^{-1}$, respectively, over the Bay of Bengal in April–May; not shown). Interestingly, the notable difference in the magnitude of the changes in all-sky and clear-sky shortwave heating (in the presence of a decrease in specific humidity) indicates the important role of cloud adjustments: a reduction of middle and high clouds allows more radiation to reach the lower troposphere, while simultaneously an increase of low clouds (due to higher relative humidity in the more stable boundary layer) leads to additional upward shortwave radiation.

[12] Aerosol-induced changes are placed in a large-scale context with the help of Figure 3. The April pattern of the trend in surface temperature is characterized by substantial cooling centered over northwestern India (Figure 3a). Associated with increased sea level pressure over the continent (Figure 3a), low-level northeasterlies blow across the Indian peninsula reinforcing the climatological wind (Figure 3d). During May, anomalous subsidence associated with reduced precipitation results in sea level pressure

increase over the Bay of Bengal and the appearance of a large-scale low-level anticyclonic circulation bringing warm moisture-laden air to the northwest (Figure 3e). The springtime warming of SST north of the equator is important for the monsoon annual northward migration toward the continent. Increased surface westerlies over the northern Bay of Bengal are likely responsible for the local surface cooling (Figure 3h), given its collocation with positive evaporation changes. Along the north-equatorial Indian Ocean, the May changes in the wind weaken the climatological westerlies, thereby further reducing evaporation (Figure 3h). Simultaneously, surface solar radiation increases due to reduced aerosol forcing (Figure 2) as well as decreased cloudiness. These factors all contribute to increasing the SST, though the ocean's inertia delays the relative warming to late May/early June (Figures 3h and 3i and from a pentad analysis here not shown), while the change is almost muted in the May mean. Interestingly, the pattern of the May minus April precipitation change is remarkably similar (with opposite sign) to the climatological variation (Figure S6): the May precipitation reduction can thus be interpreted as inhibition of its climatological northward shift. Warm advection is the major contributor to the May relative warming of the northwestern regions (not shown). Associated with the temperature increase, sea level pressure decreases, the circulation reverses (Figure 3f), moisture transport toward central India reinforces the climatological mean, and, also facilitated by the SST relative warming, precipitation moves northward and increases in June (Figure 1).

[13] The land thermodynamic forcing is further emphasized by the changes in the subcloud moist static energy (MSE, Figure S7): during April–June, anthropogenic aerosols cause a substantial increase in MSE over northwestern India and Pakistan and, by reducing humidity, a decrease over the eastern regions, the area of large climatological MSE. This northwestward anomalous MSE gradient acts to move convection inland [e.g., *Privé and Plumb*, 2007; *Wang et al*., 2009]. The northward advection of climatological large MSE by the anomalous circulation plays the major role in the development of the continental MSE anomaly in May. On the other hand, the model simulates a minor increase in heating by aerosol absorption of shortwave radiation (maximum of $\sim 0.04\ \mathrm{K\ d^{-1}\ (50\ yr)^{-1}}$) in April over northwestern India, associated to a modest increase of absorbing aerosols AOD (Figure S8). This additional heating does not lead to enhanced vertical motion in April, despite the atmospheric fast adjustment time scale. Thus, atmospheric heating by shortwave absorption appears to play a secondary role in driving the monsoon migration, though its positive trend in spring helps in setting favorable regional conditions.

4. Conclusions

[14] This work aimed at a better understanding of the mechanism driving long-term changes in the onset of the Indian summer monsoon. May and June precipitation underwent a remarkable increase over central and northern India during 1950–1999, leading to an earlier onset by about 10–20 days . This feature is in stark contrast with the substantial precipitation decrease during July–September, which dominated the summertime mean drying.

[15] Historical experiments with a state-of-the-art global climate model showed that the earlier onset was most likely a two-step radiative-dynamical adjustment to increased anthropogenic aerosols and subsequent induced thermodynamic changes in the spatial distribution of monsoon heating sources. Aerosols, mostly from intense biomass burning, exert a strong forcing over the Bay of Bengal and Indochina in spring, preventing precipitation to advance northward by increasing the atmospheric stability and cooling the SST. The atmospheric circulation adjustment results in a large near-surface thermodynamic perturbation over northwestern India, which drives the stronger precipitation northwestward migration during June. Modest changes in local shortwave heating by absorbing aerosols over the land are found to play a secondary role in establishing favorable conditions for the monsoon advancement.

[16] The mechanism described is suggested to be initiated by the direct radiative effect of aerosols dispersed over the eastern Bay of Bengal in early spring. This is not at odd with the findings of *Bollasina et al*. [2011] on the predominance of the aerosol indirect effect for the summertime precipitation decrease over India. The total aerosol optical depth over India tends to reach a peak value of about 0.8 in June and remains as high as 0.6 during summer despite some washout by monsoon rains [e.g., *Ganguly et al*. 2012; P. Ginoux, personal communication, 2013]. Additionally, indirect effects are strongest when clouds and precipitation are at their peaks. We also speculate that this effect might be further enhanced by the cloudiness increase in June (Figure 1). The unavailability of long-term in situ AOD observations as well as uncertainties/improvements in aerosol emissions are two shortcomings that need to be acknowledged and further addressed by carefully designed sensitivity studies. In this regard, anthropogenic sources of dust (e.g., those modulated by human changes in land use) represent a potential important contribution to the total aerosol forcing over India [*Ginoux et al*., 2012] to be accounted for in models.

[17] A major finding of this study is the importance of aerosol-precipitation-circulation interactions, an issue which has not been fully addressed in the past but is emerging as a major uncertainty in the understanding of the aerosol impact, especially at regional scale [e.g., *Wang*, 2013]. Additionally, we underline the importance of coupled air-sea interactions in determining the May-to-June changes. This is further supported by the analysis of AM3 historical simulations forced by observed SSTs and aerosol emissions. Interestingly, these experiments are not able to capture the precipitation increase over India in June, and precipitation is essentially reduced through all the summer months (not shown). Additionally, fast atmospheric processes alone (isolated by a similar experiment with preindustrial SSTs and aerosol emissions only evolving) would partially offset the June negative precipitation trend due to the slow response of the climate system (i.e., due to SST changes alone), further highlighting the importance of fast atmospheric adjustments. As there are no comparable studies available, the robustness of our results needs to be tested using other models, especially with respect to the relative role of local versus remote aerosol forcing [e.g., *Teng et al*., 2012; *Mahmood and Li*, 2013b].

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